THE RETRIEVAL OF SNOW GRAIN SIZE FROM SPACE USING MERIS AND AATSR TOP-OF-ATMOSPHERE SPECTRAL REFLECTANCE MEASUREMENTS

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ABSTRACT

A newly developed algorithm for the determination of snow grain size using spaceborne observations is described in detail. The algorithm was implemented in the BEAM satellite data processing software for MERIS and AATSR. The applied method includes a comprehensive cloud screening and ice/snow detection procedures.

1. INTRODUCTION

The changes of cryosphere have profound climatic consequences because of their contribution to global warming of the planet. Therefore, the task to monitor snow and ice parameters using spaceborne observations is of special importance. The main parameters of interest are as follows: snow cover, snow pollution, snow depth, snow grain size (SGS), snow water equivalent, snow temperature, snow and ice albedo, ice concentration, ice extent and thickness. In this work we present selected results related to monitoring of snow cover parameters derived with MERIS data in the framework of the ESA project “Snow_Radiance”. In particular, a newly developed algorithm for the determination of snow grain size is described in detail. The applied method includes a comprehensive cloud screening and ice/snow detection procedures. The ultimate objective of “Snow_Radiance” Project is the derivation / validation of snow algorithms for the future Sentinel 2 and 3 missions.

The retrieval code is based on the asymptotic solution of the radiative transfer equation valid for weakly absorbing multiply light scattering media. This enables the fast runs, which is essential element of operational procedures. The results of retrievals are accompanied by validation using the ground measurements.

Understanding global physical properties of snow and also trends in snow cover and pollution is of a great importance for a number of disciplines including climate studies, environmental physics and snow hydrology (Dozier, 1987; Massom et al., 2001). In this paper we address a question of subsurface snow grain size monitoring using optical measurements. It is known that the snow grain size determines the level of light absorbance by snow and this parameter is needed to asses the heat balance in snow, and also timing and magnitude of snowmelt.

The retrieval of SGS using MODIS data have been performed by Tedesco and Kokhanovsky (2007) and also by Lyapustin et al. (2009) using a similar approach as described here. The earlier version of the algorithm and applications to GLI retrievals are given by Zege et al. (1998, 2008). In this paper we concentrate on retrievals using MERIS/ENVISAT. The retrievals of snow grain size using AATSR are described by Kokhanovsky and Schreier (2009).

2. RETRIEVAL ALGORITHM

The developed retrieval algorithm for the SGS determination is based on the look-up-table (LUT) approach. In particular, the Fourier components of the reflection function in the visible (for a nonabsorbing snow) are tabulated using the code developed by Mishchenko et al. (1999). The code solves the Ambartsumian nonlinear integral equation for the harmonics $R^m(\mu,\mu_0)$ of the reflection function. These harmonics are stored in LUTs. Then the reflection function at any relative azimuth angle φ is found as

$$R(\mu,\mu_0,\varphi)=R^0(\mu,\mu_0)+2\sum_{m=1}^{M_{\max}}R^m(\mu,\mu_0)\cos(m\varphi). \quad (1)$$

Here $\mu_0=\cos\vartheta_0$, $\mu=\cos\vartheta$, ϑ_0 and ϑ are solar and viewing zenith angles, respectively. The value of $M_{\max}$ is chosen from the condition that the next term does not contribute more than 0.01% in the sum (1). In principle one more dimension (for a given phase function) in this LUT is needed and this is the dimension of the single scattering albedo. Taking into account that MERIS measurements stops at the wavelength $1\mu m$ and ice is only weakly absorbing in the spectral range of MERIS ($0.4-1\mu m$), we use the asymptotic radiative transfer theory for calculations of snow reflectance at absorbing wavelengths. This also simplifies the retrieval algorithm reducing it to analytical equations. Therefore, no

minimization procedure is required.

In particular, we use the following representation valid as $\omega_0 \to 1$ (Zege et al., 1991; Kokhanovsky and Zege, 2004):

$$R(\mu, \mu_0, \varphi) = R_0(\mu, \mu_0, \varphi) A^{f(\mu, \mu_0, \varphi)}, \qquad (2)$$

where

$$A = \exp\left\{-4s/\sqrt{3}\right\},\ s = \sqrt{\frac{1-\omega_0}{1-g\omega_0}},$$

$$f = \frac{u(\mu_0)u(\mu)}{R_0^{-1}(\mu, \mu_0, \varphi)},\ u(\mu) = \frac{3}{7}(1+2\mu).$$

Here R_0 is the reflection function of a semi-infinite snow layer under assumption that the single scattering albedo is equal to one, ω_0 is the single scattering albedo, g is the asymmetry parameter, and A is the spherical snow albedo. R_0 is calculated using Eq. (1).

The only approximation as compared to the exact RT calculations involved is the use of the term A^f in Eq. (2) to characterize light absorption by snow. We found that errors are below 6% as compared to exact radiative transfer calculations at the wavelengths 0.52-1.24 μm and $\vartheta_0 = 54^\circ$ for all azimuthal angles and $\vartheta \le 75^\circ$. Namely these short wavelengths will be used here for the inverse problem solution. In the case of MERIS wavelengths 443 and 865nm, the errors are smaller than 2% at ϑ<40 degrees typical for MERIS observations. This is well inside the calibration error of MERIS. If high accuracy is of not primary concern than an approximation for the function $R_0(\mu, \mu_0, \varphi)$ given in Appendix can be used. This speeds up retrievals and make it easier to perform various types of sensitivity studies.

MERIS does not have channels above 1 μm and, therefore, the approximation proposed here is very relevant to the interpretation of MERIS observations over snow fields. This is due to the fact that the snow albedo (and the accuracy of the approximation) increases for shorter wavelengths. The forward model itself and also errors of atmospheric correction introduce much larger errors as compared to differences between approximate and exact theories.

Eq. (2) can be used for the analytical determination of ω_0 and, therefore, a_{ef} from the snow reflection function measurements. As a matter of fact in case of small grains and the MERIS wavelengths, even simpler approximation can be used. This approximation follows from Eq. (2) as $\omega_0 \to 1$:

$$R(\mu, \mu_0, \varphi) = R_0(\mu, \mu_0, \varphi) - \frac{4s}{\sqrt{3}} u(\mu) u(\mu_0). \qquad (3)$$

Eq. (3) also enables the determination of the snow spectral albedo:

$$A(\lambda) = (R_{mes}(\lambda)/R_0)^{1/f} \qquad (4)$$

from measurements of the spectral reflection function just at one observation geometry. It is assumed that the atmospheric correction has already been performed and the influence of atmosphere is removed from the value of $R_{mes}(\lambda)$. The determination of the snow albedo also means that the snow reflection function $R = \pi I / \mu_0 E_0$ (E_0 is the solar irradiance) and the snow bi-directional reflection function BRDF=R/π are also determined simultaneously at any viewing geometry. Also the spectral snow similarity parameter is determined (see Eq. (2)):

$$s(\lambda) = \frac{\sqrt{3}}{4} \ln\left[\frac{1}{A(\lambda)}\right]. \qquad (5)$$

This parameter is of importance for understanding of radiative transfer in snow.

The technique given above enables the determination of spectral characteristics $A(\lambda), s(\lambda), BRDF(\lambda)$ without a priori assumptions on the size of grains up to the wavelength 1.24 μm. We also found that a particular nonspherical grain shape assumption is not crucial for the snow albedo retrieval. In fact different assumptions on the grain shape produce very similar values of the spectral albedo.

The single scattering albedo can be found from the expression for the similarity parameter if the value of the asymmetry parameter is known. At least for dry snow, one can assume that the asymmetry parameter $g = 0.76$ independently on the wavelength (in the spectral range considered) as for fractal grains (Kokhanovsky and Zege, 2004). For the wet snow, the value of g increases and the retrieval results for the single scattering albedo (but not the snow surface albedo) will be biased. One derives:

$$\omega_0(\lambda) = \frac{1-s^2(\lambda)}{1-gs^2(\lambda)}. \qquad (6)$$

An important point is that although ω_0 will be possibly biased due to the assumption on the value of the asymmetry parameter, the spectral behavior of ω_0 is not effected by this assumption because (for large snow grains) g is almost spectrally neutral parameter. Moreover, because $gs^2 \to 0$ in the spectral range studied, the influence of the incorrect assumption on the value of the asymmetry parameter does not influence results for ω_0 considerably. In many applications not ω_0 but rather the probability of photon absorption (PPA) $\beta = 1 - \omega_0$ is needed. It follows for this parameter:

$$\beta(\lambda) = \frac{(1-g)s^2(\lambda)}{1-gs^2(\lambda)} \qquad (7)$$

and the error in $\varepsilon = 1 - g$ influences results

considerably. However again, the spectrum $\beta(\lambda)$ is not much effected by the assumption on the value of ε because it is almost a spectrally neutral parameter in the spectral range considered.

The shape of particles must be assumed for the retrieval of the grain effective size a_{ef} from the value of PPA given by Eq. (7). For this one can use the following exponential approximation (EA) derived from fitting geometrical optics calculations (Kokhanovsky and Nauss, 2005): $\beta = \beta_\infty(1-\exp(-\alpha\ell))$, where $\alpha = 4\pi\chi/\lambda$, $\beta_\infty = 0.47$, $\ell = Ka_{ef}$, λ is the wavelength, χ is the imaginary part of the ice refractive index. The parameter *K* depends on the shape of particles and the value of *K*=2.63 can be used for fractal snow grains (Kokhanovsky and Nauss, 2005).

The value of the single scattering albedo in the near infrared ($\lambda \geq$ 0.8-1.0 μm) is almost independent on the snow pollution. Therefore, it is proposed to find a_{ef} in the near infrared (e.g., 1.02 μm or at 0.865 μm). Then the retrieved value of the grain effective radius can be used to determine the fraction of the PPA related to the pollution (in the visible). Actually, if one chooses the wavelength of 443nm, then the imaginary part of the refractive index of ice is so small ($\sim 10^{-10}$) that the whole absorption can be attributed to impurities and not to snow grains.

At the wavelength 0.865 μm, there is a chance (for a highly polluted snow only) that the signal is contaminated by the contribution of pollutants. This contamination effect can be easily accounted for slightly modifying the algorithm described above.

Namely, we use the fact that it is possible to write for channels 1(0.443 μm) and 2(0.865 μm) in the approximation under study:

$$R_1 = R_0 \exp(-\gamma\sqrt{\beta_1}), \tag{8}$$

$$R_2 = R_0 \exp(-\gamma\sqrt{\beta_2}), \tag{9}$$

where indices 1 and 2 signify the channel,

$$\gamma = \frac{4f}{\sqrt{3(1-g\omega_0)}}. \tag{10}$$

We will neglect the difference of ω_0 from 1.0 in the dominator of Eq. (10).

Here we assume that there is some light absorption by snow even in the visible (e.g., due to soot). The value of probability of photon absorption can be written as

$$\beta = \frac{N_i C_{abs,i} + N_s C_{abs,s}}{N_i C_{ext,i} + N_s C_{ext,s}}. \tag{11}$$

Here

$$N_s = \frac{c_s}{\bar{V}_s} \tag{12}$$

is the number concentration of soot particles, $\bar{V}_s$ is their average volume, c_s is the volumetric concentration of soot(the fraction of volume filled by soot), $C_{abs,\alpha}$ is the average absorption cross section of soot particles, $C_{ext,s}$ is the average extinction cross section of soot particles. Parameters with the index "i" have the same meaning as described above except for ice.

We will neglect the contribution of soot to the general light extinction in snow. Then it follows:

$$\beta = \beta_i + \beta_s, \tag{13}$$

where $\beta_i = \frac{C_{abs,i}}{C_{ext,i}}$ is given by EA and

$$\beta_s = \frac{V_i c_s C_{abs,s}}{V_s c_i C_{ext,i}}. \tag{14}$$

The average extinction cross section of the ice grains $C_{ext,i}$ can be estimated as follows:

$$C_{ext,i} = \frac{\bar{\Sigma}_i}{2}. \tag{15}$$

Here $\bar{\Sigma}_i$ is the average surface area of grains. Taking into account that $C_{ext,i}/\bar{V}_i = 1.5\, a_{ef}^{-1}$ (Kokhanovsky and Zege, 2004) in this approximation and also assuming that $C_{abs,s}/\bar{V}_s = \varsigma\alpha_s$, which is true in the Rayleigh domain for small soot particles ($\varsigma = 0.84$ at the soot refractive index $n = 1.75$ (van de Hulst, 1957), $\alpha_s = 4\pi\chi_s/\lambda$, $\chi_s = 0.46$), we derive:

$$\beta_s = \frac{2}{3}\varsigma c\alpha_s a_{ef} \tag{16}$$

where

$$c = c_s/c_i \tag{17}$$

is the relative soot concentration. The mass absorption coefficient of soot $\sigma_{abs} = C_{abs}/\rho_s\bar{V}_s$ is equal to $\varsigma\alpha_s/\rho_s$ in the considered approximation. Here ρ_s is the soot density. Assuming that $\varsigma = 0.84$, $\chi_s = 0.46$, $\lambda = 443nm$, $\rho_s = 1g/cm^3$, one derives: $\sigma_{abs} = 8.4g/m^2$, which is close to the modern estimates of this parameter ($7.5 \pm 1.2m^2/g$ (Bond and Bergstrom, 2006; Flanner et al., 2007).

Therefore, we can write:

$$R_1 = R_0 \exp\left[-\gamma\sqrt{\frac{2}{3}\varsigma\alpha_{s,1}ca_{ef}}\right], \tag{18}$$

$$R_2 = R_0 \exp(-\gamma\sqrt{\beta_{i,2} + \frac{2}{3}\varsigma\alpha_{s,2}ca_{ef}}). \tag{19}$$

Here we neglect the light absorption processes in ice at the first wavelength. These two equations can be used to find both the size of ice crystals and the concentration of pollutants. It follows from the first equation for X= ca_{ef} :

$$X = \frac{3}{2\varsigma\gamma^2\alpha_{s,1}}\ln^2 r_1 \qquad (20)$$

and, therefore,

$$\beta_{i,2} = \frac{\ln^2 r_2}{\gamma^2} - \frac{2}{3}\varsigma X\alpha_{s,2}\,. \qquad (21)$$

Here we introduced the normalized reflectance: $r_i \equiv R_i / R_0$. The value of a_{ef} can be found from equations given above:

$$a_{ef} = K\alpha_{i,2}^{-1}\ln\left[\frac{\beta_\infty}{\beta_\infty - \beta_2}\right]. \qquad (22)$$

Then the concentration of soot is determined as $c = X / a_{ef}$. In practice, one measures the concentration of soot as the fraction of soot mass in a given mass of snow $c_f = c_s\rho_s / c_i\rho_i$, where ρ_s is the density of soot and ρ_i is the density of ice. Therefore, for the transformation of the satellite – derived c to the ground measured values of c_f, one must use the multiplier $\eta = \rho_s / \rho_i$:

$$c_f = \eta c\,. \qquad (23)$$

We will assume that $\eta \equiv 1$ in this study. It is known that $\rho_i = 0.917 g/cm^3$. The density of soot depends on its structure. It varies in the range 1-2 g/cm^3. The assumption of $\eta \equiv 1$ is consistent with the lower limit of this variability.

For MODIS, the channel at 1.24 μm is available in addition to 0.865 μm channel. Generally, the wavelength 1.24 μm is the best for retrievals in the case of a homogeneous snow because then even heavy pollution does not influence the results of the grain size retrieval (therefore, one can put *X*=0 in the expression for β_2 and derive the following simplified equation: $\beta_2 = \gamma^{-2}\ln^2 r_2$, which can be used for the retrievals of a_{ef}). For vertically inhomogeneous snow, this wavelength brings information only from the top of the layer and may be not sensitive to the grains at deeper layers seen by the 443nm wavelength used for the snow pollution retrieval. Even if measurements at 865nm are used, there is quite large a mismatch in the volume of snow sensed using 443nm and 865nm wavelengths. We found that the Jacobians for the soot concentration (at 443nm) approach zero at the distance of 20cm from the top layer and the values of Jacobians for the SGS vanish already at 2-5cm depending on the wavelength. Therefore, possible soot layer deposited at, say, 5cm from the snow top will influence the signal in the visible but not at 865nm. This makes application of dual-wavelength algorithm not possible in this case and one should use the single channel algorithm. This comment is relevant to other multiple-wavelength snow retrieval algorithms as well (Zege et al., 2008; Lyapustin et al., 2009).

It follows in the case of multiple pollutants:

$$\beta = \frac{N_i C_{abs,i} + \sum_{\alpha=1}^{M} N_\alpha C_{abs,\alpha}}{N_i C_{ext,i} + \sum_{\alpha=1}^{M} N_\alpha C_{ext,\alpha}}\quad. \qquad (24)$$

Here

$$N_\alpha = \frac{c_\alpha}{V_\alpha} \qquad (25)$$

is the number concentration of α-pollutant particles, V_α is their average volume, c_α is the volumetric concentration of α-pollutant(the fraction of volume filled by this particular pollutant), $C_{abs,\alpha}$ is the absorption cross section of the α-pollutant, $C_{ext,\alpha}$ is the extinction cross section of the α-pollutant. If dust is present in large quantities, the second term in the dominator of equation for β can not be ignored and such parameters as the size of dust grains and also their concentration must be determined along with the parameters for soot. The necessity for such retrievals occurs only at rare occasions (heavy dust pollution events) and we will not explore this opportunity in this work.

3. SNOW GRAIN SIZE RETRIEVALS USING MERIS ONBOARD ENVISAT

Let us apply the method derived to the MERIS data. We have selected the clear sky scene over Greenland for the retrievals of the snow grain size using MERIS data. After the application of cloud screening algorithm and the use of the SGS retrieval algorithm we derived the SGS distribution as shown in Fig.1. The histogram of the retrieved SGS is given in Fig.2. The results are realistic and found values of a_{ef} are in the expected limits. Unfortunately, there were no grain size observations on the ground. Therefore, the validation of retrievals using MERIS data presented here is not possible.

However, it was possible to validate our algorithm using MODIS data. The results of validation are reported in Fig.3. The ground measurements have been performed by MRI (Japan) team led by T. Aoki. We found a clear correlation of the satellite-derived and ground-measured snow grain sizes. The part of the discrepancy is due to the fact that the snow grain size measured on the ground lucks the repeatability and also the snow reflectance measured on a satellite poorly correlates (see Fig.4) with the ground-derived grain size, which indicates that the snow grain size measured on the ground does not correspond to the effective optical grain size defined via

ratio of the grain volume to its surface area ($a_{ef} = 3V/\Sigma$ as used in this paper for the SGS).

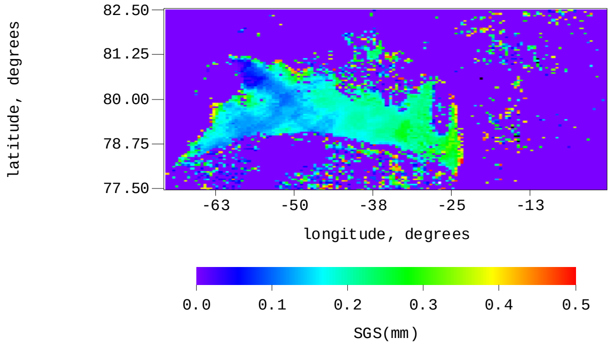

Fig.1. The retrieved snow grain size spatial distribution. The MERIS measurements over Greenland have been performed on June 21, 2004 (ENVISAT orbit 12075, 15:42:49).

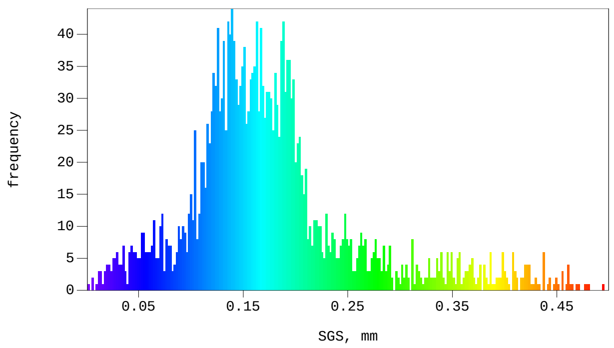

Fig.2. The retrieved snow grain size frequency distribution.

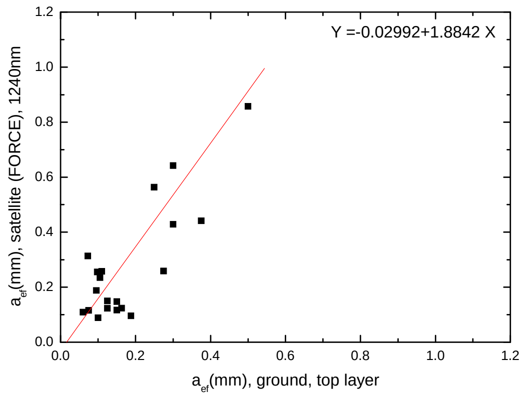

Fig.3. The correlation of MODIS-derived and in situ SGS.

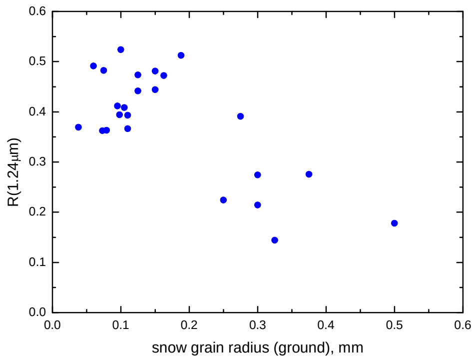

Fig.4. The correlation of the MODIS-measured snow reflectance and the snow grain size measured *in situ*.

4. CONCLUSIONS

In this work a new snow grain size retrieval algorithm based on MERIS measurements is described. The algorithm is applied to MERIS data and validated using ground snow size measurements. The discrepancies found are related to the fact that the optical snow grain size as measured using satellite data is defined in a different way as compared to the grain size measured on ground (the width of the narrowest part of the crystal). The spatial variation of the SGS can also contribute to the discrepancy between in situ and satellite-derived SGS. This is due to the fact that the satellite measurements analyzed have a spatial resolution of $1\,km^2$ and ground measurements are performed just in several points inside the MERIS ground scene. This together with some other factors can explain the differences shown in Fig.3. For the future validation activities, it is desirable to make measurements of SGS on ground using optical measurements (e.g., the spectal reflectance) and not manual observations of snow crystals.

5. APPENDIX. SIMPLE APPROXIMATION FOR THE REFLECTION FUNCTION

The reflection function $R_0(\mu,\mu_0,\varphi)$ can be calculated in the following approximation (Kokhanovsky, 2006):

$$R_0(\mu,\mu_0,\varphi) = \frac{A + B(\mu+\mu_0) + C\mu\mu_0 + p(\theta)}{4(\mu+\mu_0)},$$

where *A*=1.247, *B*=1.186, *C*=5.157, $p(\theta) = 11.1\exp(-0.087\theta) + 1.1\exp(-0.014\theta)$, θ is the scattering angle. This angle is given in degrees here and defined as $\theta = \mathrm{acos}(-\mu\mu_0 + ss_0\cos\varphi)$, φ is the relative azimuth, $\mu = \cos\vartheta$, $\mu_0 = \cos\vartheta_0$, $s = \sin\vartheta$, $s_0 = \sin\vartheta_0$. The accuracy of this approximation is better than 15% for the MERIS observation conditions (Kokhanovsky, 2006).

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